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Climate response to basin-specific changes in latitudinal temperature gradients and implications for sea ice variability

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Abstract. We use experiments with the GISS general circulation model to investigate how changes in latitudinal temperature gradients affect atmospheric circulation in different ocean basins, with particular attention paid to the implications for high-latitude sea ice. The results are relevant to both estimated past climate changes, current climate gradient changes (e.g., El Niño–Southern Oscillation events), and proposed future climate responses to greenhouse gases. Sea surface temperature gradients are increased/decreased in all ocean basins, and in the Pacific and Atlantic separately, without changing sea ice or global average temperature. Additional experiments prescribe sea ice growth/reduction with global cooling/warming. As expected, increased gradients strengthen the subtropical jet stream and deepen the subpolar lows in each hemisphere, but results in the Northern and Southern Hemispheres differ in fundamental ways. In the Northern Hemisphere, increased storm intensities occur in the ocean basin with the increased gradient; in the Southern Hemisphere the deeper storms occur in the ocean basin with the decreased gradient. Alterations of the gradient in one ocean basin change longitudinal temperature gradients; an increased gradient in one basin from tropical heating results in subsidence in the tropics in the other basin, mimicking the effect of a decreased gradient in that basin. The subtropical jet is therefore strengthened over the basin with the increased gradient and decreased over the other ocean basin. Hence in many respects, regional effects, such as the strength of subpolar lows in an individual basin, are amplified when the gradient changes are of opposite sign in the two ocean basins. The Southern Hemisphere response occurs because gradient increases in one ocean basin, by strengthening the subtropical jet, shift storm tracks equatorward and away from the potential energy source associated with cold air advection from Antarctica. At the same time, with a weaker subtropical jet in the other basin, storms move poleward and strengthen. This latter effect may explain observed sea ice variations that are out of phase in the Atlantic and Pacific Ocean basins in the Southern Hemisphere (referred to as the Antarctic dipole) as well as upper ocean variability in the Weddell gyre. Gradient changes produce little effect on sea level pressure in the Arctic, unless sea ice is changed. With Arctic sea ice reductions, the sea ice response acts as a positive feedback, inducing cyclonic circulation changes that would enhance its removal, as may be occurring due to the current high phase of the North Atlantic Oscillation.

1. Introduction

The recent changes in Northern Hemisphere sea ice coverage and thickness [*Parkinson et al.*, 1999; *Rothrock et al.*, 1999; *Johannessen et al.*, 1999] have raised the issue of whether they are indeed trends or part of natural variability patterns. One of the most robust projections of general circulation models (GCMs) in response to increasing anthropogenic trace gases is a decrease in Northern Hemisphere sea ice. Is what we are currently seeing, the decrease of 2.8%/decade over the 40 years and the noticeable thinning of ice throughout the region, the expected beginning of this process? If so, it has strong implications for high-latitude warming and a positive sea ice/albedo feedback.

Alternatively, the sea ice changes could be the response to the increasing frequency of the high phase of the Arctic Oscillation (AO) (or the closely related North Atlantic Oscillation) [*Hurrell*, 1995; *Walsh et al.*, 1996; *Thompson and Wallace*, 1998] characterized by lower pressure over the pole. The cyclonic circulation associated with this change could be leading to greater export of ice through the Fram Strait [*Mysak et al.*, 1996; *Kwok and Rothrock*, 1999; *Polyakov et al.*, 1999; *Kwok*, 2000; *Hilmer and Jung*, 2000; *Deser et al.*, 2000], reducing the Arctic ice cover. This circulation can also alter the surface freshwater distribution leading to a loss of the Arctic cold halocline layer and commensurate initiation of significant ocean heat flux resulting in considerable winter ice thinning [*Martinson and Steele*, 2000]. This would then be a dynamic forcing, rather than the primarily thermodynamic forcing described above.

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Even if this were the cause, it would not necessarily indicate whether it was the result of natural or anthropogenic forcing. The increased high phase of the AO might simply represent natural variability in the system, and be reversed at any moment, or it might represent forcing associated with greenhouse gases. In one mechanism, *Shindell et al*. [1999] suggested that the AO high phase arose through the response of the system to increased greenhouse gases, via the impact on the temperature gradient and zonal wind structure. Along a constant pressure surface intersecting the tropopause, greenhouse gas increases induce tropospheric warming at lower latitudes and stratospheric cooling at higher latitudes. This gradient change results in stronger zonal winds in the lower stratosphere, forcing more equatorward planetary wave propagation. Poleward angular momentum transport associated with this change in wave refraction strengthens the zonal winds in the lower stratosphere, and the subsequent effect on planetary waves in the troposphere allows the effect to extend down to the surface. In this scenario, and in the model, the effect is likely to continue. Observations suggest this mechanism may indeed be operating [*Baldwin and Dunkerton*, 1999; *Thompson and Wallace*, 2000].

Another possible relationship of sea ice changes to global warming similarly emphasizes the increased high phase of the North Atlantic Oscillation but ascribes the high phase to changes in North Atlantic heat transports [*Fyfe et al*., 1999; *Russell et al*., 2000]. Anthropogenic greenhouse-initiated warming of the tropical oceans, associated with the shallow tropical mixed layer depths, produces more moisture in the atmosphere. This added moisture is subsequently advected poleward, and in the North Atlantic results in increased rainfall, which freshens the ocean. North Atlantic Deep Water (NADW) production and associated oceanic poleward heat transport is inhibited, cooling the North Atlantic. The increased temperature gradient in the Atlantic is then associated with stronger west winds. Evaporation decreases despite the wind speed increase [*Russell and Rind*, 1999], associated with the cooler ocean temperatures, amplifying the freshening. Observations of a shift in tendency for convection and NADW production from the Norwegian to the Labrador Seas [*Dickson et al*., 1996] might be part of this pattern, and cooler surface air temperatures have been observed over the North Atlantic region [e.g., *Hansen et al*., 1999]. *Vinnikov et al*. [1999] concluded that the decrease in Arctic sea ice is most likely a consequence of greenhouse warming, with a small chance that it could occur from natural variability, based on the results of simulations with the Geophysical Fluid Dynamics Laboratory coupled atmosphere-ocean model.

While one mechanism favors forcing from above the troposphere and the other forcing from below, what they have in common is a change in the latitudinal temperature gradient associated with tropical warming and extratropical cooling. This raises the more general question of the response of circulation features at high latitudes to altered latitudinal temperature gradients, both globally and in various ocean basins. If the North Atlantic cools due to the NADW suppression, global warming could lead to a locally increased temperature gradient in that basin. If NADW production does not change, then the expected high-latitude amplification of climate forcing should result in a decreased latitudinal temperature gradient.

In the North Pacific, standard general circulation model (GCM) future scenarios for CO_2 increase, such as those given in *International Panel on Climate Change (IPCC)* [1995, chap. 6], show a reduction in temperature gradient, again due to high-latitude amplification processes such as sea ice reduction. However, if El Niño frequencies were to increase, or become permanent as in some GCM scenarios, then the Pacific temperature gradient might well increase. The high-latitude circulation systems would presumably respond differently in these different cases, with various combinations possible, as would be the effect on sea ice advection and sea ice change. Furthermore, the result would be interactive: the more sea ice reduction, the more of an increase in latitudinal temperature gradient likely.

What about in the Southern Hemisphere? *Stammerjohn and Smith* [1997] and *Yuan and Martinson* [2000] have shown that variations occurring in sea ice coverage are out of phase between the eastern Pacific and the Atlantic. *Yuan and Martinson* [2000] suggest that this response is related to El Niño–Southern Oscillation (ENSO) [*Carleton*, 1988; *Simmonds and Jacka*, 1995; *Ledley and Huang*, 1997]. Moreover, *Yuan et al*. [1999], in a case study of storm activities in late 1996 based on space-based observations of surface winds, found that anomalous cyclonic activity had increased in the South Pacific and decreased in the South Atlantic. Could it be the result of a latitudinal temperature gradient change in the South Pacific due to the La Niña conditions that existed in 1996? Likewise, D. G. Martinson and R. A. Iannuzzi (Spatial/temporal patterns in Weddell gyre characteristics and their relationship to global climate, submitted to *Journal of Geophysical Research*, 2000) (hereinafter referred to as MI2000) find that the upper ocean characteristics of the Weddell gyre covary with the sea ice and ENSO in a manner they hypothesize reflects enhanced winter cyclonic forcing during El Niño periods and diminished cyclonic forcing during La Niña. Can we understand why this would cause an alternation in sea ice and polar sea ice changes in the different ocean basins?

The experiments described below are used to examine these questions in the context of a general circulation model. We shall show that altered latitudinal gradients will affect pressures at high latitudes, in ways that are unique to the different ocean basins, and differ in the Northern and Southern Hemispheres.

2. Experiments

The simulations employ the latest version of the GISS GCM (model II′), which has been well documented in the literature at the $4° \times 5°$ resolution [*Rind and Lerner*, 1996; *Hansen et al*., 1997; *Rind*, 1998]. The model is reasonably representative of the state of the art for a coarse resolution model. In GCM radiation comparisons with line-by-line calculations [*Cess et al*., 1993] the GISS model was among the most accurate. This was also true in an intercomparison of cloudiness and its variability in a large number of GCMs [*Weare et al*., 1995], and in intercomparison of seasonal changes in cloud radiative forcing in GCMs [*Cess et al*., 1997]. A comparison of the hydrologic processes in 29 GCMs with observations ranks the GISS model in the upper quartile [*Lau et al*., 1996]. A comparison among 30 GCMs of the amplitude and seasonality of precipitation over the United States showed the GISS model to be one of the most realistic, with a fidelity similar to that of the finer resolution 19-layer T42 model of the Max Planck Institute [*Boyle*, 1998].

Our intent is to provide a general discussion of the types of processes that arise when gradients are changed, either when the gradient is changed uniformly or when changed in one

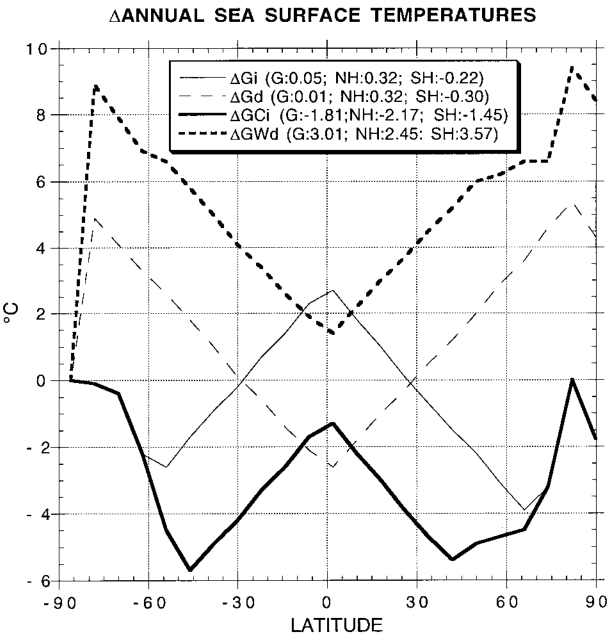

Figure 1. Annual average sea surface temperature anomalies used in experiments Gi, Gd, GCi, and Gwd. Global, Northern Hemisphere, and Southern Hemisphere anomaly values are indicated in parentheses. The anomalies in Gi and Gd were used in all the other experiments when changing sea surface temperature gradients in individual ocean basins.

ocean basin relative to another. In this sense we are not attempting to simulate any particular climate, past or future, though some features of our experiments may be similar to characteristics found in some paleoclimate and future climate scenarios, as well as in ENSO-related variability. The gradient changes we use are therefore generic. Some of the experiments were described by *Rind* [1998] (henceforth paper 1), and the rest were introduced briefly by *Rind* [2000]. The sea surface temperature gradients in the GCM were changed by increasing (decreasing) tropical temperatures by 3°C and decreasing (increasing) high-latitude temperatures by 6°C, with a linear change in between, as shown in Figure 1 (Gi and Gd) (the figure also shows the global, Northern Hemisphere (NH) and Southern Hemisphere (SH) average sea surface temperature (SST) anomalies). This SST gradient change was chosen to be sufficiently large to produce a clear signal. To put it in perspective, it is associated with a surface air temperature gradient change between the equator and 60°N of ±3.3°C, similar to that associated with $2 \times CO_2$ experiments, about one-half that estimated to occur in the Last Glacial Maximum (in the Atlantic) and about double that in typical El Niños (in the Pacific). The procedure was chosen to limit the overall global mean surface air temperature change, which would otherwise provide effects in addition to that associated with the altered gradients. Were the planet all ocean, the conservation would be exact; as it was, the global mean surface air temperature showed only small changes (of the order of 0.1°–0.2°C [*Rind*, 1998]). To separate the effects of sea ice changes on the gradient, in most of the experiments sea ice coverage is kept fixed at current day values, so the practical effects of the SST changes resulting from exposure to the atmosphere (e.g., the surface air temperature responses) do not extend beyond about 70° latitude. In several additional experiments (GCi and GWd in Figure 1, explained below) we explore the effects of changing sea ice as well, in which case significant air temperature responses extend to the North Pole.

Table 1. Description of Experiments

Label	Description
Gi	increased gradient in all ocean basins
Gd	decreased gradient in all ocean basins
GCi	increased gradient in all basins plus 4°C cooling and increased sea ice
GWd	decreased gradient in all basins plus 4°C warming and reduced sea ice
Ai	increased gradient in Atlantic Ocean
Pi	increased gradient in Pacific Ocean
Ad	decreased gradient in Atlantic Ocean
Pd	decreased gradient in Pacific Ocean
AiPd	increased gradient in Atlantic, decreased gradient in Pacific
AdPi	decreased gradient in Atlantic, increased gradient in Pacific

The following experiments were performed (Table 1). Using the same magnitude gradient changes as described above for paper 1, in Experiment Gi, the SST latitudinal gradient was increased uniformly in both hemispheres in all ocean basins, while in Experiment Gd, it was decreased. In Experiment "Ai" the latitudinal temperature gradient was increased only in the Atlantic Ocean, all global ocean points that fall between 20°E and 95°W, with a northern boundary at 72°N. In Experiment "Pi" the gradient was increased only in the Pacific Ocean, all global ocean points between 65°W and 120°E. Similar experiments were also conducted with decreasing the gradients in the individual ocean basins (Ad, Pd, respectively). In a seventh experiment, the gradient was increased in the Atlantic and decreased in the Pacific (AdPi), mimicking results from some increased greenhouse gas experiments. Its inverse was also run (AiPd) for comparison. In Experiment GCi a uniformly increased latitudinal gradient change was combined with a 4°C uniform temperature reduction, and sea ice prescribed to increase wherever the SSTs dropped below the freezing point of ocean water (−1.56°C in the model); and in Experiment GWd, a uniformly decreased latitudinal gradient change was combined with a 4°C uniform increase in temperature, and sea ice was prescribed to decrease wherever appropriate. The sea surface temperature changes in these last two experiments are also shown in Figure 1; the deviations from linear gradients at the higher latitudes are due to the effects on the zonal average SST of the prescribed sea ice changes. The effect on high-latitude pressure systems in these experiments is a result of both the altered global temperatures and the sea ice changes. A complete list of the experiments is provided in Table 1.

In addition to changing the latitudinal gradient, altering sea surface temperatures in individual ocean basins also changes the interbasin longitudinal gradients. This has the most direct effect in the Antarctic circumpolar region, where the Pacific and Atlantic waters are contiguous. To minimize abrupt longitudinal shifts in that region, we modify the sea surface temperatures between 60° and 70°W to produce smooth gradients. Additional modified transitions occur near the borders of the Indian Ocean with the Atlantic and Pacific.

Each experiment was run for 6 years, with the results averaged over the final five years. A comparison with 15-year simulations shows that this is sufficient to establish the dynamic differences between simulations with changes as large as these, when sea surface temperatures are specified. The one excep-

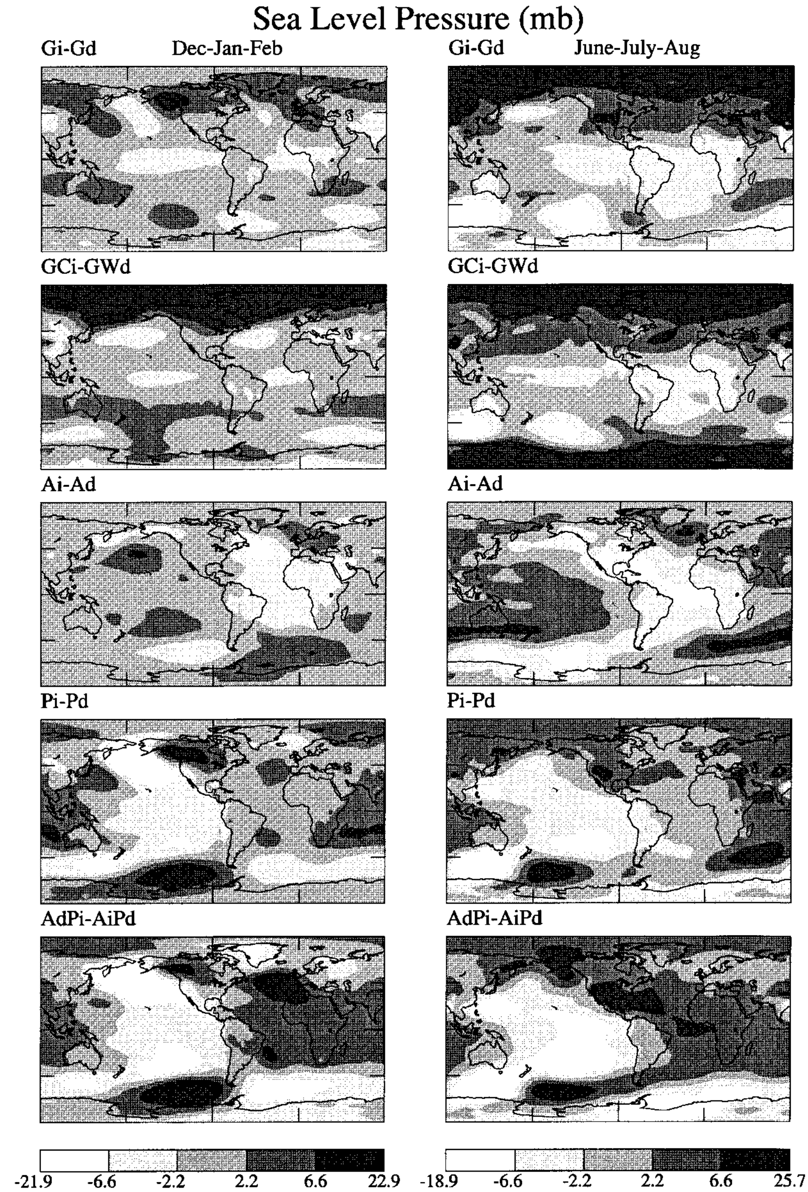

Figure 2. Sea level pressure differences between the different experiments. Shown are the changes between experiments Gi and Gd (top, row 1), GCi and GWd (row 2), Ai and Ad (row 3), Pi and Pd (row 4), and AdPi and AiPd (bottom, row 5). Results are given for December through February on the left and June through August on the right. The sea level pressure differences are 5 year averages.

tion is the storm track diagnostics, which are accumulated for the last 10 years of 12-year simulations. Significance levels are provided for the changes in the different parameters, as derived from interannual variations of a 12-year control run with specified SSTs, and the discussion is limited to those differences of greater than 95% significance.

3. Results

3.1. Sea Level Pressures

Since changes in cyclonicity are the main interest in this context, we show first the variation in sea level pressure, given in Figure 2, for the solstice seasons. To emphasize the effects, we have differenced experiments with opposite gradient changes (henceforth referred to as inverse experiments). At high latitudes, changes greater than 6.6 mbar (the extreme shades) are generally significant at 95%, while at low latitudes, changes greater than 1 mbar are similarly significant. As in paper 1, the results are consistent with expectations in the Northern Hemisphere: an increased gradient and larger available potential energy lead to eddy energy increases. Thus during winter the Aleutian Low is strengthened when the Pacific gradient is magnified (with a Pacific-North American-(PNA) type teleconnection pattern generated), as shown in the fourth panel on the left-hand side of Figure 1; the Icelandic Low in a southwest position is stronger with the amplified Atlantic gradient (third panel, left-hand side). With either ocean basin experiencing an increased gradient, the other ocean basin experiences some pressure rises, reminiscent of the third eigenvector in the 500-mbar height field normalized covariance matrix [*Wallace and Gutzler*, 1981]. Overall, the North Pacific sea level pressure field seems more responsive to gradient changes

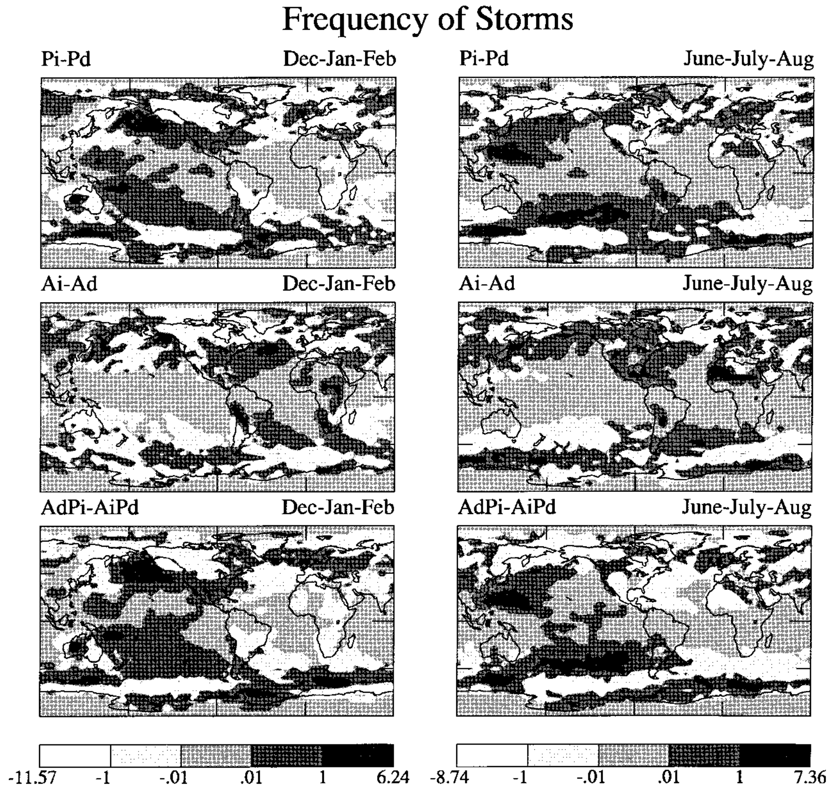

Figure 3. Change in storm frequency for Pi-Pd (top), Ai-Ad (middle), and AdPi-AiPd (bottom) for December–February (left) and June–August (right). Storms are defined on the sea level pressure maps as closed lows with mean pressures less than 1000 mbar. Results are averaged over the last 10 years of 12 year simulations.

than the North Atlantic when sea ice is invariant, due, perhaps, to the greater size of the Pacific Ocean and hence the latitudinal gradient change, as well as the greater sea ice in the North Atlantic limiting the effect of the sea surface temperature change. Given the "opposite response" in the single ocean basin experiments, a linear expectation would be that the effect on the subpolar lows would be even greater when gradient changes are of opposite sign in the different ocean basins; this is indeed the case; for example, the Aleutian Low is deeper and broader in AdPi-AiPd (fifth panel, left-hand side) than it is in Pi-Pd. We return to this point below.

Over the Arctic the increased temperature gradient without any sea ice change (Gi − Gd) produces little sea level pressure response in winter. However, when sea ice is increased (and decreased in its inverse experiment), sea level pressure is considerably higher (lower) (GCi − GWd). The lower pressure over the pole in GWd will have an additional effect not included here; since increased cyclonic flow leads to a reduction in Arctic sea ice cover, it will have a positive feedback by exposing more open water, decreasing atmospheric stability, and leading to even further increases in cyclonic behavior. During summer an increased gradient by itself is sufficient to result in higher sea level pressures, perhaps because colder local conditions (and greater vertical stability) dominate the pressure response when horizontal temperature gradients are normally weaker, as in summer.

In the Southern Hemisphere, uniformly increased gradients result in deeper subpolar lows, with and without sea ice changes. During winter the increased low-pressure regions are in the South Pacific, while during summer, they are primarily in the Atlantic/Indian Ocean sector. This greater cyclonicity even occurs when sea ice is increased, in contrast to the situation in the Arctic. (As the Arctic is at higher latitudes than the subpolar positions being discussed for the Southern Hemisphere, it has much more sea ice to begin with, and so when sea ice is not changed, it experiences less of a temperature gradient change and less sea level pressure response.)

However, in the Southern Hemisphere the results for the experiments in which the gradient is changed separately in individual ocean basins provide a very different picture. Now when the gradient increases in one ocean basin, the subpolar low in that basin actually decreases in intensity (higher sea level pressure), while the subpolar low in the other ocean basin strengthens (lower sea level pressure). This is completely opposite to the situation in the Northern Hemisphere. We explain below why this occurred.

3.2. Storm Tracks and Intensities

To complement the change in standing wave patterns associated with the seasonal average sea level pressure field, we show, in Figure 3, the change in the frequency of storms between the inverse experiments with basin-specific changes and, in Figures 4a and 4b, the mean intensity of those storms in December–February and June–August, respectively (not shown as differences, because were storms not to be present in a grid box in one of the simulations, the difference in intensity would be meaningless). For the purposes of increasing the significance and confidence in the results, the experiments and control were run for 12 years for these diagnostics; in general, frequency changes in the extreme categories shown in Figure 3

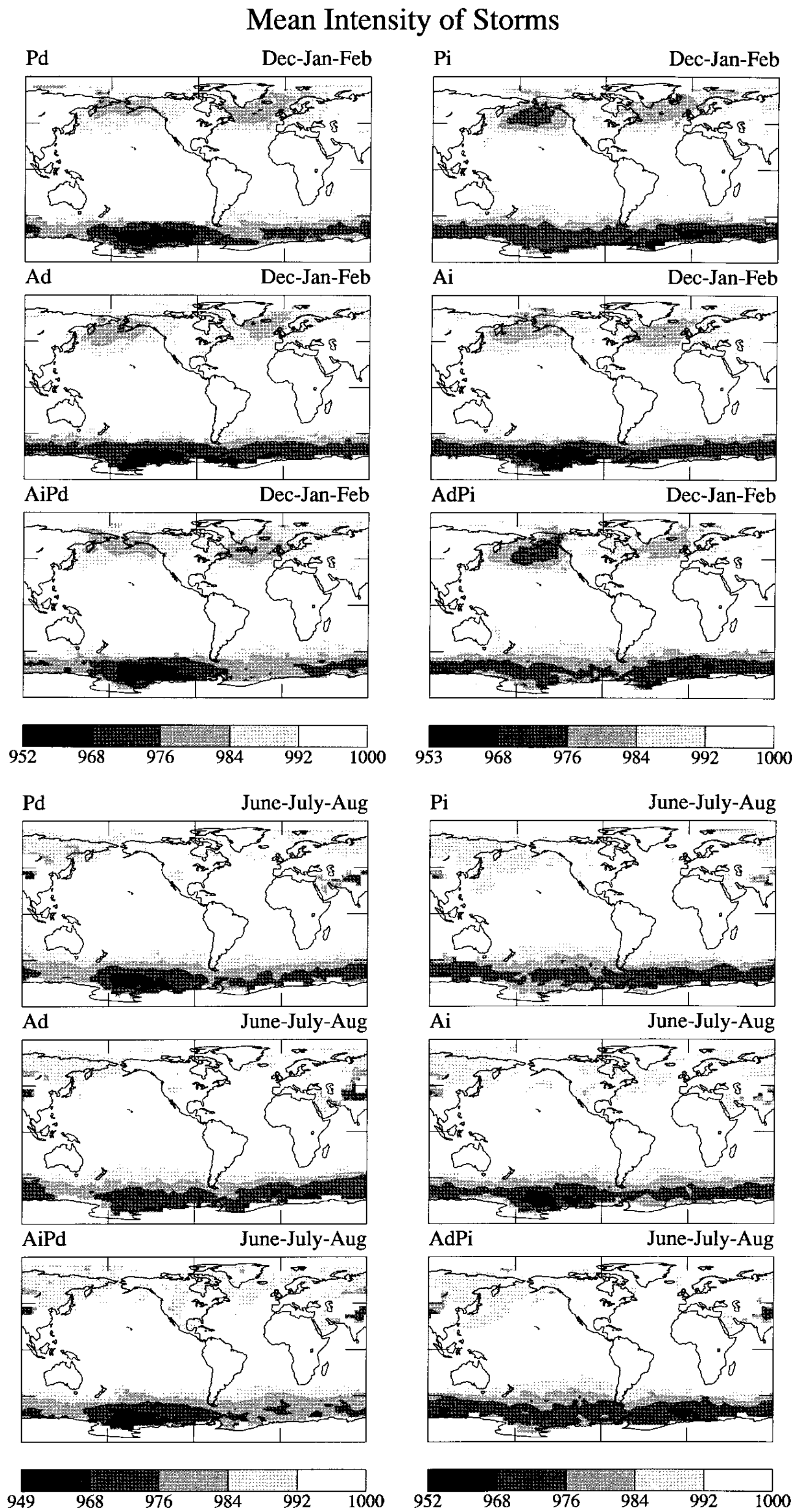

Figure 4. Average storm intensity (central pressure) for the experiments with ocean-specific gradient changes. (a) December–February, (b) June–August. Results are averaged over the last 10 years of 12 year simulations.

are significant at the 95% level, as are mean intensity differences of 10–15 mbar. Storms are defined on the basis of their sea level pressure as closed lows with mean pressures less than 1000 mbar. Basing the storm definition on its sea level pressure minimum with a closed contour has its drawbacks, particularly in the Southern Hemisphere where it tends to emphasize slower and deeper cyclones in contrast to more mobile and open depressions [e.g., *Sinclair*, 1994; *Simmonds and Keay*, 2000]. Other definitions are available for storm definition and intensity, such as the maximum of the Laplacian of the pressure, although this tends to produce a scale selectivity. Additional considerations are involved in choosing a single limiting value (1000 mbar) for the cutoff; the variation of the background pressure with latitude would imply that the same deviation from latitude-mean conditions might qualify as a storm at higher latitudes but not in the subtropics; however, using deviations from the background mean then biases against cyclones at high latitudes, whose ubiquity is what lowers the latitudinal average pressure. Focusing on the differences between experiments presumably minimizes some of these problems, and the slower/deeper storms may well be more important for sea ice transports. Nevertheless, the limitations of the cyclone definition, and its intensity, should be understood.

The increased Pacific gradient is responsible for a storm track in winter similar to that often associated with El Niños, from the North Pacific across the southern United States. The Atlantic gradient affects storms primarily in the Atlantic, from the northeastern United States eastward across the Atlantic. Note that in the Southern Ocean, with gradient changes in either the Atlantic or the Pacific, the storm tracks in both seasons are farther equatorward in the ocean basin with the increased gradient and farther poleward in the other basin.

With an increased gradient, one would expect stronger storms in the respective ocean basins, and this is clearly true in the North Pacific (Figure 4a). In the North Atlantic the effect is much more muted, although the area of more intense storms is somewhat greater. In the Southern Ocean, in both seasons, the opposite effect prevails: the storms are weaker in the basin with the increased gradient as the storm track shifts equatorward, and stronger in the other basin (compare, for example, Ai and Ad or Pi and Pd in Figure 4b). Again, the effects are greatest when gradient changes are opposite in the different basins (AiPd, AdPi).

In general, then, in the Northern Hemisphere gradient changes have a somewhat stronger influence on storm tracks and storm intensities in the Pacific, and the effects of gradient changes on storm intensity in particular ocean basins in the Southern Hemisphere are opposite to those in the Northern Hemisphere.

3.3. Jet Stream Changes

An obvious influence on all transient eddy propagation is the change in the jet stream accompanying the gradient differences. Figures 5a and 5b show the changes in the 200 mbar winds for the different inverse experiments in the solstice seasons. In regions of peak wind speeds, such as the subtropics, changes of 10 ms^{-1} are significant at the 95% level, other regions in the extratropics show such significance at around 5 ms^{-1} and the rest of the tropics at 1–3 ms^{-1}. With the increased Pacific gradient, the subtropical jet is amplified across the Pacific Ocean, extending downstream over North America, as observed during El Niño events. With the decreased Pacific gradient, the jet stream core shifts poleward. The Atlantic gradient changes affect primarily the subtropical Atlantic winds, especially in winter, with some upstream effect over North America. Similar responses occur in both hemispheres. Note that in the basin-specific gradient change experiments, when the subtropical jet stream increases in one ocean basin, with the core moving equatorward, the jet stream moves poleward in the other basin, again in both hemispheres. The increased cyclonicity apparent with reduced sea ice in GWd is visible in the Arctic wind curvature even at this level (the reverse of the result shown in Figure 5a). Equatorial changes also arise, due to the longitudinal circulation effects discussed below.

3.4. Longitudinal Circulation Effects

To fully understand the results presented above, it is necessary to explore the consequences of another component of these experiments: changing gradients in individual ocean basins gives rise to a change in longitudinal gradients between the ocean basins. These gradient changes then induce longitudinal circulation cells. Various GCM simulations have been made of the effects of changes in longitudinal temperature gradients within ocean basins [e.g., *Stone and Chervin*, 1984; *Simmonds et al*., 1989]. In this case the SST gradients within each ocean basin are unchanged, but by virtue of changing the latitudinal gradient in one ocean basin only, the effect is to change the longitudinal gradient between ocean basins.

Figures 6–8 show the changes in longitudinal circulation cells and associated parameters for the three different pairs of inverse experiments. Significant changes at the 95% level occur for annual average surface wind differences of the order of 0.5 ms^{-1} and 200 mbar wind changes of 1–1.5 ms^{-1}; vertically integrated vertical velocity changes of 5–15 $\times$ 10^{-4} ms^{-1}; 320 mbar temperature changes of 0.5°C; 200 mbar relative humidity changes of 3%. When the equatorial Pacific is relatively warmer in Pi-Pd (Figure 6), increased rising air and convective heating occurs in the Pacific, and weak subsidence with little heating is induced over the tropical Atlantic (Figure 6 (middle)), a direct consequence of the altered longitudinal sea surface temperature gradients. The longitudinal circulation cell setup results in increased west winds at 200 mbar and weak easterlies at the surface, from about 90°W to 90°E (Figure 6, top). The temperature and relative humidity increase over the warm Pacific, and relative drying occurs over the Atlantic.

When the tropical Atlantic is warm in Ai-Ad (Figure 7) the situation is reversed, with the warming and rising air now over the Atlantic. Therefore when the tropical Atlantic is relatively cooler (the inverse of Figure 7), relative subsidence is occurring over the Atlantic, qualitatively similar to what happened with the warmer Pacific. Hence when the tropical Pacific is warmer and the tropical Atlantic cooler, the effects are magnified (and are to some extent additive), as shown by the results in Figure 8 for PiAd-PdAi. Changes in the longitudinal circulation cells and jet stream magnitudes arising from altered longitudinal temperature gradients are an oft-derived feature of altering longitudinal SST gradients in GCMs [e.g., *Stone and Chervin*, 1984; *Simmonds et al*., 1989].

4. Discussion

With the above results we can explain several of the more puzzling features of the sea level pressure field changes. By far the most unexpected result concerns the influence of gradients in one ocean basin on circulation elsewhere. This is particularly true with respect to the low-pressure regions surrounding Antarctica. As evident in Figure 2, the subpolar low-pressure systems in that region are actually stronger in the ocean basin that does not have the gradient increase. For example, pressures are considerably higher in the Weddell Sea when the Pacific gradient is increased, and lower (i.e., a deeper low) when the Atlantic gradient increases (Figure 2). As shown in Figure 5, an increased gradient intensifies the subtropical jet stream in the particular ocean basin and relocates storms farther equa-

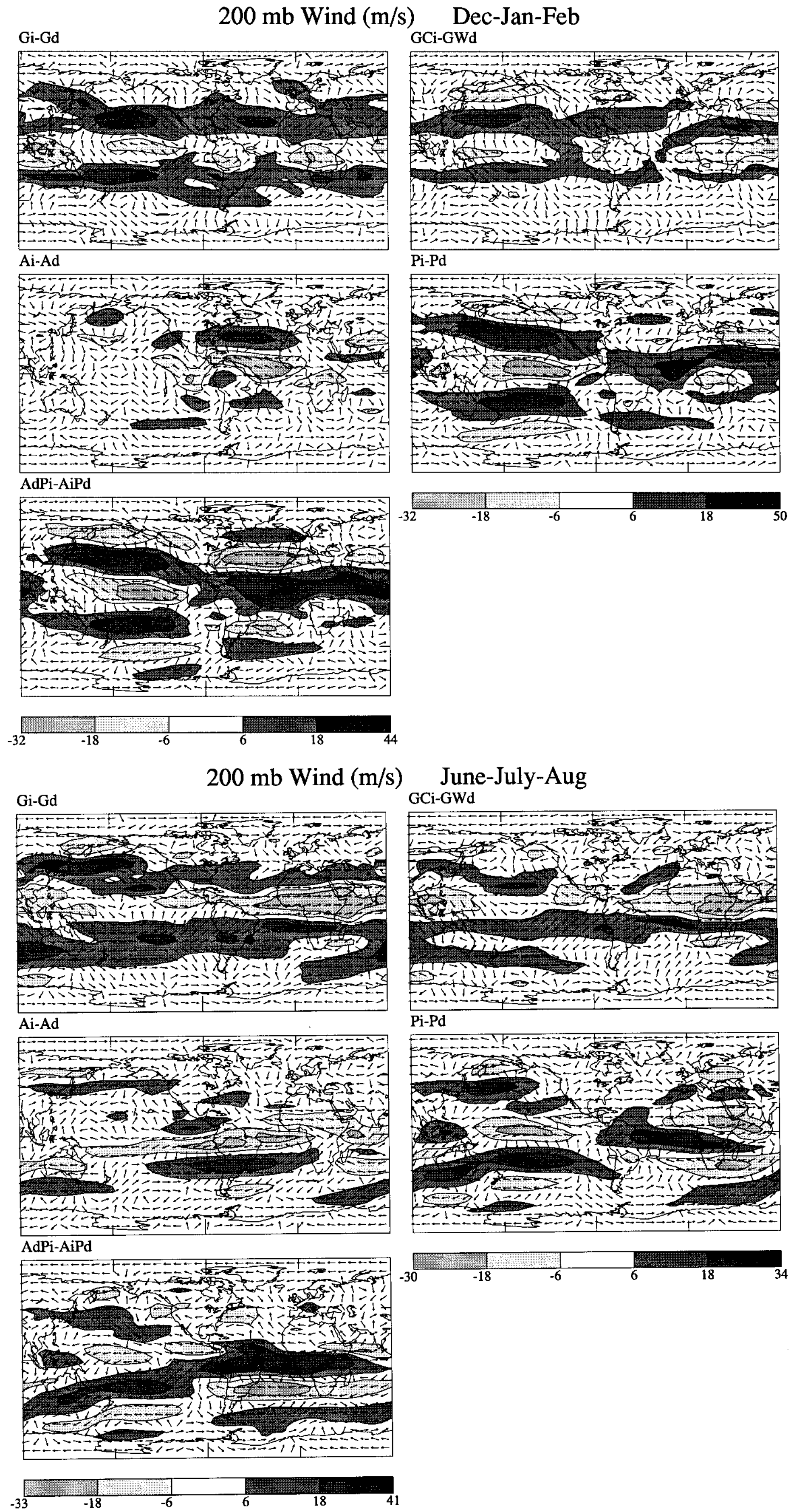

Figure 5. Wind speed changes at 200 mbar between the experiments for (a) December–February and (b) June–August. The arrows indicate the direction of wind change; the shading indicates the magnitude of the change. For clarity, only one quarter of the arrows are shown.

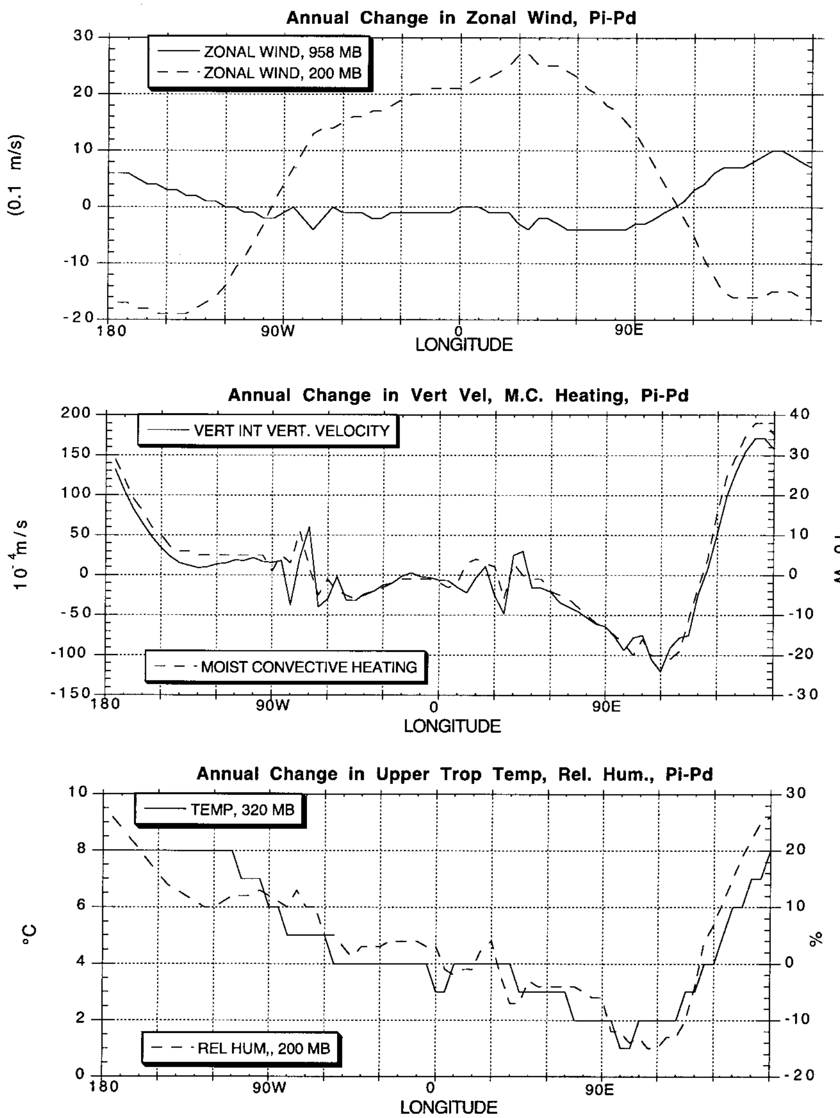

Figure 6. Tropical changes (6°N–6°S) between Pi and Pd for zonal wind at 950 and 200 mbar (top), vertically integrated vertical velocity and moist convective heating (middle), and upper tropospheric temperature and relative humidity (bottom). Note that the increased gradient runs are warmer throughout the equatorial region.

torward. In the Southern Hemisphere this has the effect of moving them farther away from Antarctica (Figure 3) and thus farther from the potential energy maximum and the strong local latitudinal temperature gradient associated with cold air coming off that continent. The result is a weakening of the storms in the ocean basin with the increased gradient and the more equatorial storm track (Figures 3 and 4).

However, why do storms increase in intensity in the other ocean basin? The results from Figures 6–8 indicate that the increased gradient in one ocean basin results in equatorial subsidence in the other basin. This subsidence is qualitatively similar to a decrease in gradient in the other ocean basin, which in the equatorial region similarly results in relative subsidence. The effect is to reduce the meridional circulation rising from the equatorial region in the ocean basin without the increased gradient and hence reduce the intensity of its associated subtropical jet stream. As noted with reference to Figure 5, the jet stream moves poleward over the ocean basin that does not have the gradient increase. Storms thus move poleward, are more likely to draw into their circulations cold air from Antarctica, and hence are more intense (Figures 3, 4). As noted by *Simmonds and Keay* [2000], Southern Hemisphere storms in winter are most intense when found immediately off the Antarctic coast.

This alternating basin effect in storm intensity does not occur in the Northern Hemisphere where storms do not depend on being close to the pole for their cold air source. Cold air advection from higher latitudes follows pathways across the North American and Eurasian continents, which keeps the air relatively unmodified. Eddy available potential energy is maximized when this cold air comes into contact with the warmer air at middle and subpolar latitudes.

Even in the Northern Hemisphere, the intensity of storms (e.g., the Aleutian Low) in an ocean basin is maximized when the gradients are of opposite sign in the two ocean basins. This is the result of the longitudinal circulation cells that arise from changes in gradients in the particular ocean basins. With the Pacific gradient increased and the Atlantic gradient decreased, rising air is somewhat intensified in the tropical Pacific (Figure 8, middle) (with subsidence in the

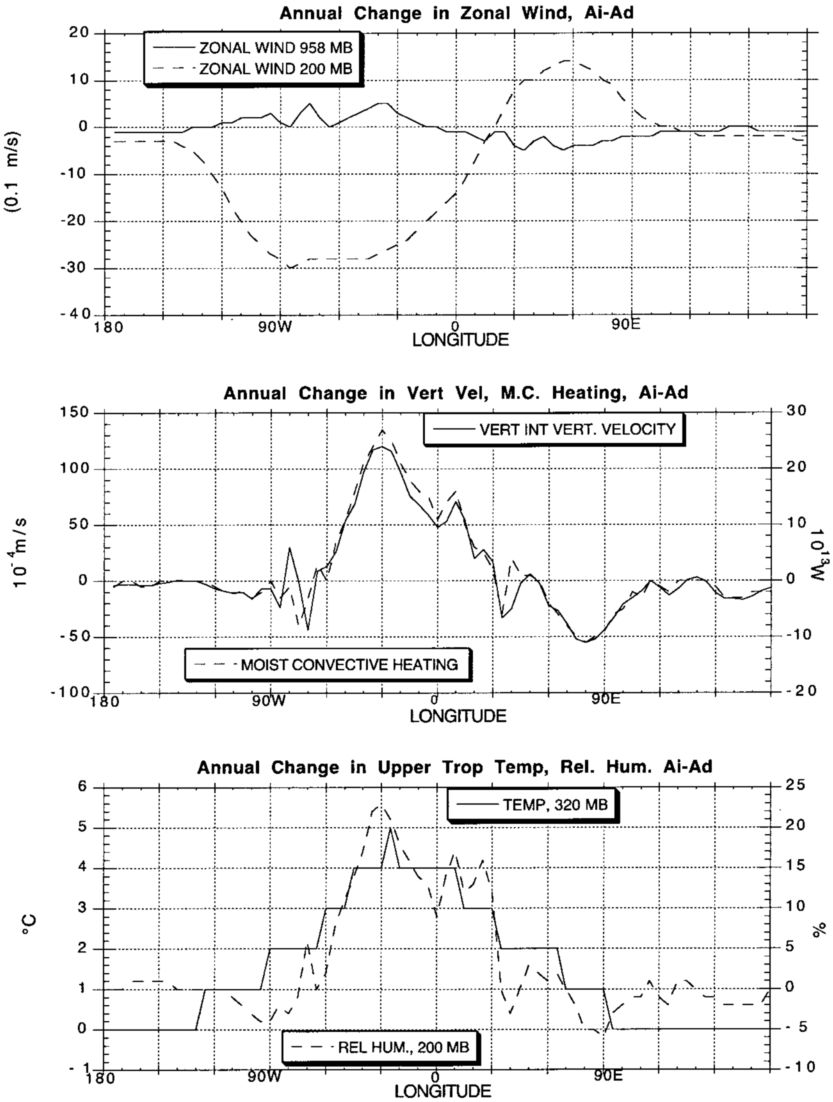

Figure 7. As in Figure 6 for Ai minus Ad.

Atlantic), as is subsidence-induced warming in the subtropical Pacific. This results in a stronger latitudinal temperature gradient and greater eddy kinetic energy at Pacific midlatitudes.

Table 2 gives the eddy energy values in the different experiments for the entire Northern Hemisphere during winter. AdPi has both the largest eddy energy and standing wave energy and the largest eddy available potential energy and eddy kinetic energy in the long waves, all features that are consistent with the greater anomalies in sea level pressure given in Figure 1.

Table 2 also shows the Southern Hemisphere winter values of eddy energy, most of which is in the transient mode (see column 7). The uniformly increased gradient experiments (Gi and GCi) have higher values than the runs with uniformly decreased gradients (Gd and GWd). The experiments with greater energy in the South Atlantic (Pi and Ad) have larger hemispheric average values than the runs with greater energy in the South Pacific (Ai and Pd). The greatest energy overall is in the combination of experiments which amplify South Atlantic eddy energy (AdPi).

5. Relevance to Observed Subpolar Changes

These results have direct relevance to the currently observed sea ice and subpolar upper ocean variations in both hemispheres. In the Northern Hemisphere the model indicates that increased SST gradients in the North Atlantic will produce a high-phase "NAO-type" response, with intensification of the Icelandic Low. However, by themselves they do not lead to decreases in sea level pressure over the Arctic (hence no AO signature). To accomplish this requires sea ice decreases, which in these experiments occurs in conjunction with a decrease in the latitudinal gradient. Hence the current increase in the AO phase would not have been forced directly by the observed increases in the Atlantic temperature gradients (associated, for example, with reduced NADW production) but could have resulted from sea ice reductions associated with the increased cyclonic flow in conjunction with the Icelandic Low intensification (high phase of the NAO). This latter effect may be the result of natural variations or of greenhouse warming, as discussed in section 1. In this sense, the NAO high phase helps generate an AO high phase through its effect on the sea ice field.

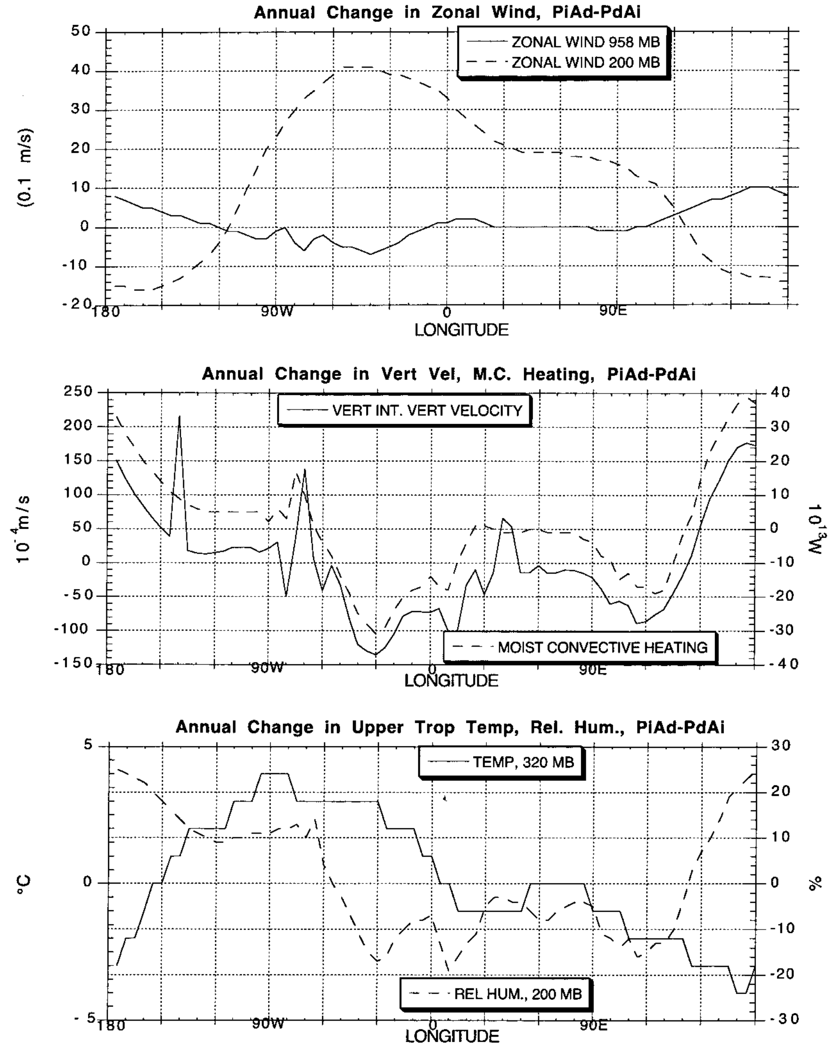

Figure 8. As in Figure 6 for AdPi-AiPd.

In the Southern Hemisphere the results suggest that alternation in cyclonicity between ocean basins would be expected if a change in latitudinal temperature gradient occurred in one of the basins, as in El Niño or La Niña occurrences. *Yuan and Martinson* [2000] do, in fact, see the effects of such alternation between the Pacific and the Atlantic in the sea ice field (their so-called Antarctic dipole), as shown in Figure 9. As an example, in late austral winter of 1996, the sea ice extent in the

Table 2. Winter Eddy Energies in Different Experiments

	NH Eddy Kinetic Energy	NH Standing Eddy Kinetic Energy	NH Transient Eddy Kinetic Energy	NH EKE Waves 1–4	NH EAPE Waves 1–4	SH Eddy Kinetic Energy
CONT	1646	510	1136	803	3460	1779
Gi	1882	543	1339	934	3411	1962
Gd	1502	510	992	755	3647	1647
GCi	1957	461	1496	932	4329	2048
GWd	1623	385	1238	850	4555	1548
Pi	2026	808	1218	1122	3665	2130
Pd	1673	552	1121	857	3516	1763
Ai	1690	588	1102	830	3456	1805
Ad	1746	493	1253	900	3619	1964
AdPi	2138	938	1200	1239	3803	2209
AiPd	1757	524	1233	910	3475	1836
s.d.	35	28	30	23	66	42

All units 10^{17} J.

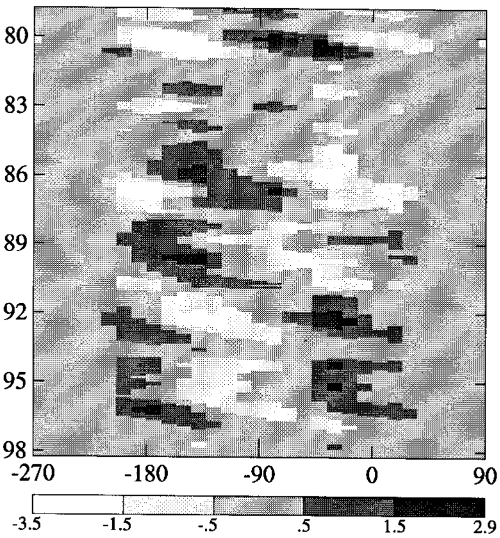

Figure 9. Southern Hemisphere sea ice edge anomaly (containing only the first two modes of variability accounting for 53% of the total variance) in the unit of degree of latitude as a function of time and longitude. A positive value indicates more extensive sea ice.

Pacific sector of the Antarctic ($-165°$ to $-195°$ in the figure) significantly exceeded the average for the 20 year time period [see also *Yuan et al.*, 1999, Figure 2]. The long-lived storms prevailed in the polar/subpolar Pacific regions. On the other hand, the Southern Atlantic ($-30°$ to $-90°$) experienced much less cyclone activity during the same period and reduced sea ice extent. Although *Yuan et al.* [1999] did not examine the variations of latitudinal temperature gradient in each basin, there is a La Niña event occurring in the tropical Pacific in 1996. The alternation of the strength of storms between the ocean basins may well be associated with the weaker latitudinal temperature gradient during the La Niña year, as suggested by and consistent with the model results. Hence we would also expect a stronger latitudinal temperature gradient, reduced storminess, and less sea ice in the South Pacific during the El Niño conditions of 1980, 1983, 1986, the early 1990s, and 1997, all years which do feature less extensive sea ice cover there (Figure 9).

The results here are also consistent with the upper ocean response observed by MI2000, whereby an El Niño event is associated with enhanced cyclonic forcing of the Weddell gyre, with the opposite effect during La Niña years.

6. Conclusions

Increasing the latitudinal temperature gradient in all ocean basins, without changing sea ice, strengthens the subpolar lows in both hemispheres but has little effect on sea level pressure over the Arctic Basin in winter. However, when sea ice is decreased, lower pressure occurs over the Northern pole as well. Hence if increased latitudinal temperature gradients in the Atlantic or Pacific result in increased advection of sea ice out of the Arctic or reduced winter ice growth, the reduced sea ice cover will probably result in lower atmospheric pressure, which would further increase cyclonic wind flow and sea ice advection. Whatever is responsible for the increased phase of the NAO may therefore be producing similar effects via this positive feedback between the AO and the sea ice reduction.

In the Southern Hemisphere, increasing the latitudinal temperature gradient in only one ocean basin results in decreased storm intensity in that basin. This occurs as a result of the intensified subtropical jet stream, which accompanies the increased gradient, redirecting storm tracks equatorward, moving them farther from the potential energy source associated with the cold air coming off Antarctica. At the same time, an increased gradient in one ocean basin in the Southern Hemisphere results in increased storm intensity in the other ocean basin. This effect is explained by the fact that an increase in the latitudinal gradient in one ocean changes the longitudinal temperature contrast with the other ocean basin. Consequently, warming of the tropical Pacific results in relative subsidence over the tropical Atlantic, an effect which mimics, to some extent, a decrease in the Atlantic Ocean latitudinal temperature gradient (which also produces subsidence in the tropical Atlantic).

These results are consistent with recent observations that show increasing the temperature gradient in the Pacific (in association with El Niño events) weakens the subtropical jet stream in the Atlantic (e.g., *Boyle* [1996, Figure 14], shown for both observations and GCMs, although the effect is not remarked upon). As indicated in these experiments, the jet stream weakening allows for a poleward shift of the storm tracks and leads to the generation of stronger storms in the ocean basin without the latitudinal gradient change, and vice versa. Variations in sea ice coverage, observed to be out of phase in the two ocean basins in the Southern Hemisphere, may arise from this process and accompany ENSO-induced latitudinal temperature gradient changes.

These results can be of use in interpreting paleoclimate observations (as discussed by *Rind* [2000]), of special importance in trying to deduce the tropical temperature changes that have actually occurred over time. In the current and future climate the Northern Hemisphere sea ice change can produce a positive feedback by amplifying wind changes that are helping to produce the sea ice response. In addition, for future climate considerations the amplifying effect of opposing changes in latitudinal temperature gradients in different ocean basins could magnify atmospheric circulation and sea ice changes, were such opposing gradient changes to occur as suggested in section 1. This amplification of regional effects is true for other parameters as well, such as rainfall over land areas, a result that will be explored in a subsequent paper.

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